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A THEORETICAL ESTIMATE OF THE AVERAGE VERTICAL DISTRIBUTION OF TEMPERATURE IN THE MARTIAN ATMOSPHERE

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ABSTRACT

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The average variation of temperature with height in the Martian atmosphere is probably controlled by radiative and convective processes. With the use of a simple theoretical formulation in which it is assumed that convection will extend to that height above which the radiative equilibrium lapse rate is just stable, the average temperature profile is computed. It is assumed that the average surface temperature is 230K, there is no absorption of solar radiation in the atmosphere, and carbon dioxide in an amount equal to 2% by volume is the only important radiating gas. The radiation fluxes are computed with the aid of radiation tables; the radiative equilibrium temperatures are calculated using an iterative procedure. The computed temperature profile is characterized by an adiabatic troposphere extending to about 9 km, above which the temperature continues to decrease with height to an average value of about 90K for the topmost 5mb layer.

AUTHOR

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I. INTRODUCTION

The average vertical distribution of temperature in a planetary atmosphere is generally controlled by the following processes: (1) radiation, (2) convection, and (3) condensation phenomena. Horizontal eddy transport and advection of heat, though important at various latitudes and seasons, should not affect the average vertical temperature distribtuion. The Martian atmosphere and, to a smaller extent, the earth's atmosphere are largely transparent to solar radiation. Thus, most of the solar energy is not absorbed directly in the atmospheres of these planets but rather at the planetary surfaces. This energy is then transmitted to the atmosphere by infrared radiative and convective transfer. In the earth's atmosphere, but probably not on Mars, the release of heat into the atmosphere by condensation of water vapor accompanies the radiative and convective processes. The combination of these processes in the earth's atmosphere results in an average tropospheric lapse rate of about 6K/km, which is about 60% of the lapse rate that would be produced by dry convection alone: the adiabatic lapse rate. Radiative processes, both short-wave absorption and long-wave emission, if allowed to act alone in the earth's atmosphere would produce a temperature distribution close to adiabatic in the troposphere and close to the observed temperatures in the upper stratosphere (Manabe and Möller, 1961). Thus, in the earth's stratosphere, the controlling processes are principally radiative.

The temperature distribution in the Martian atmosphere is unknown. Present knowledge concerning the Martian atmosphere indicates that condensation phenomena and absorption of solar radiation by the atmosphere can be neglected. The major processes affecting the average vertical temperature distribution in the Martian atmosphere are then infrared radiation and convection. In the present study, we compute the average vertical temperature distribution on Mars on the basis of a simple model which, though primarily a radiative equilibrium model, permits the effect of convection to be considered.

II. THEORETICAL FORMULATION

As stated in the Introduction, the important processes controlling the temperature structure of the Martian atmosphere are infrared radiation and convection. If one computes the temperature distribution that would result from radiative transfer alone, one obtains a radiative equilibrium temperature profile. For an atmosphere such as Mars' or the earth's, the infrared radiative equilibrium temperature profile is characterized by extreme convective instability in the lower layers (see, for example, Goody, 1954, De Vaucouleurs, 1954). Because of its instability, such a temperature profile cannot exist; convection takes place and the temperatures in the lower layers are changed. The new lapse rate is determined by convective equilibrium rather than radiative equilibrium and is equal to the adiabatic lapse rate (in the case of the earth's atmosphere condensation processes further change the prevailing lapse rate). The height to which convection extends depends upon infrared radiation processes; convection will extend to that height above which infrared radiation produces a stable lapse rate (Goody, 1954). This level separates a convective troposphere from a radiative equilibrium stratosphere and can be referred to as the tropopause. Simple as this theoretical formulation may be, it does improve upon a theory based upon radiative equilibrium alone in that the effect of convection, if not the actual mechanisms, are included, and it does permit estimates of the height of the tropopause. This formulation has been applied to the Martian atmosphere previously

(Goody, 1957); however, the surface temperature used, 270K, is probably about 40K higher than the average surface temperature (see section on physical model).

III. COMPUTATIONAL TECHNIQUE

Our goal is to determine the average temperature structure in the Martian atmosphere on the basis of a convective troposphere lying below a radiative equilibrium stratosphere that is thermally stable. The computations are based upon the initial value method of computing radiative equilibrium temperatures (see Manabe and Möller, 1961). Starting with an initial isothermal atmosphere whose temperature is equal to the average Martian surface temperature, we compute the infrared radiative temperature change rates as a function of height. These rates are then applied to the initial temperature profile in order to obtain a new temperature profile. This process is continued until a temperature profile is obtained for which the radiational rates of temperature change are negligibly small. At each stage of the calculations, however, the temperature profiles are checked for instability; if any layer has a lapse rate greater than the adiabatic, the temperatures are brought back to the adiabatic curve prior to the next calculation of radiational rates of temperature change. Such instabilities will occur, according to the previous discussion, in the lower layers of the atmosphere. Throughout the computation the surface temperature is held constant. Within the limits of the theoretical formulation, the final temperature profile obtained from these calculations should be representative of the average temperature structure of the Martian atmosphere.

The radiational rates of temperature change were determined from the divergence of the net flux of radiation with height, as computed with the aid of Elsasser's (1960) radiation tables. Carbon dioxide is assumed to be the only important radiating gas, and the 15μ band of carbon dioxide the only important band. Following Elsasser's notation, the upward and downward fluxes of radiation at any reference level can be written as (see Elsasser, 1960, and Hales, et al, 1960)

$$F\uparrow\ (0) = \sigma T_g^{\,4} - \int_{T_o}^{T_g} R(u,T)\,dT \tag{1}$$

$$F\downarrow\ (0) = \int_{T_1}^{T_o} R(u,T)\,dT + \int_{o}^{T_1} R(u_1,T)\,dT \tag{2}$$

where

$$R(u,T) = \int \frac{dB_\nu}{dT}\left[1 - \tau(u)\right]\,d\nu \tag{3}$$

and $F\uparrow$ (0) and $F\downarrow$ (0) are the upward and downward radiation fluxes at the reference level, σ is the Stefan-Boltzmam constant, T_g is the ground temperature, u is the reduced carbon dioxide path length (increasing upwards and downwards from the reference level), T is temperature, T_1 is the temperature at the top of the atmosphere, T_o is the temperature at the reference level, u_1 is the reduced carbon dioxide path length from the top of the atmosphere to the reference level, B_ν is the black-body energy per unit spectral interval, and τ is the flux transmissivity. The integral in Equation (1) and the first integral in Equation (2) follow the actual temperature-path-length relationship in atmosphere; the second integral in Equation (2) is a boundary term; the integral in Equation (3) extends over the 15μ carbon dioxide band.

The net flux at any level is given by the difference between the upward and downward fluxes.

$$F_{net}(0) = F\uparrow\ (0) - F\downarrow\ (0) \tag{4}$$

The radiational rate of temperature change can be obtained from the vertical divergence of the net flux

$$\frac{\Delta T}{\Delta t} = \frac{g}{c_p} \frac{\Delta F_{net}}{\Delta p} \tag{5}$$

where $\Delta T/\Delta t$ is the radiational rate of temperature change for a layer of pressure thickness Δp, g is the acceleration of gravity, c_p is the specific heat of the atmosphere at constant pressure. With $g = 373$ cm sec^{-2}, $c_p = 0.24$ cal g^{-1}deg^{-1}, ΔF_{net} in cal cm^{-2}day^{-1} and Δp in mb, Equation (5) can be written as

$$\frac{\Delta T}{\Delta t}\left(\frac{deg}{day}\right) = 1.55 \frac{\Delta F_{net}}{\Delta p} \tag{6}$$

The atmospheric integrals in Equations (1) and (2) are evaluated numerically by dividing the atmosphere into eight layers of thickness 10 mb and one layer (the topmost) of 5 mb. Mean temperatures and path lengths for each layer are used to obtain the values of R from expanded tables of R, which have been computed by Hales, et al (1960). As suggested by Elsasser (1960), a linear pressure correction is applied to the path lengths to correct for pressure broadening of the absorption lines.

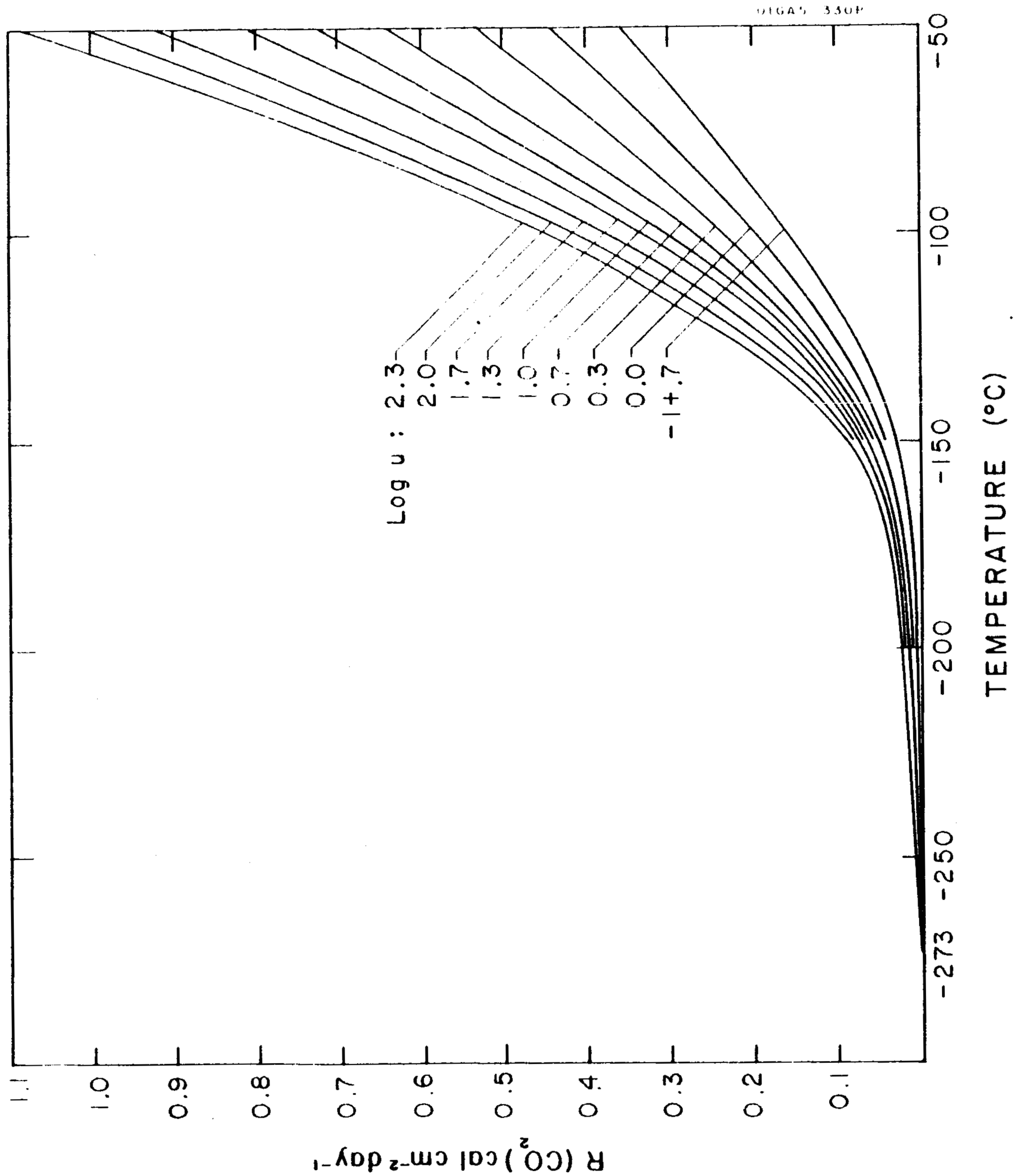

Figure 1. Extrapolated values of Elsasser's R function for carbon dioxide at selected path lengths.

IV. PHYSICAL MODEL

A value of 230K is adopted for the average surface temperature of Mars. This value is based upon a theoretical estimate (Ohring, et al, 1962) that is in reasonable agreement, after allowing for diurnal effects, with the indications of surface temperature obtained from observations of the thermal emission of Mars. The most probable value of surface pressure is about 85 mb (De Vaucouleurs, 1954) and the adiabatic lapse rate is about 3.7K/km (Kellogg and Sagan, 1961).

The exact composition of the Martian atmosphere is still uncertain, but nitrogen is believed to be the major component. Water vapor has not been detected spectroscopically but is believed to be present in small quantities (less than about 10^{-2} cm precipitable water). Ozone has not been detected but may be present; an upper limit of 0.05 cm has been given by Kuiper (1952). In the present calculations, the Martian atmosphere is assumed to be dry and devoid of ozone. The only gas that has been detected spectroscopically in the Martian atmosphere is carbon dioxide. Grandjean and Goody (1955) in a re-analysis of Kuiper's (1952) near infrared measurements derive a carbon dioxide content of about 2% , assuming a surface pressure of 85 mb. This compares to an average carbon dioxide content of 0.03% by volume in the earth's atmosphere. Assuming that the carbon dioxide is uniformly mixed in the Martian atmosphere, we can write the pressure corrected path length as

$$\Delta u = 43.4 \left(\frac{\bar{p}}{1000} \right) \Delta p \qquad (7)$$

where Δu is the carbon dioxide path length in cm NTP of a layer of thickness Δp mb and average pressure $\bar{p}$ mb. With this physical model and with the theoretical formulation and computational techniques outlined above, we can now compute the average vertical distribution of temperature in the Martian atmosphere.

V. RESULTS

The computed temperature profile and an adiabatic profile are illustrated in Figure 2; the crosses represent temperatures obtained from an atmosphere that is initially isothermal at 230K, and the circles represent temperatures obtained from an atmosphere that initially has an adiabatic temperature profile and a surface temperature of 230K. At no level is the difference between the two sets of computed temperatures greater than 5K. The troposphere extends to a height of about 9km where it is topped by a tropopause whose temperature is 196K. Above the tropopause the temperature decreases at an average rate slightly less than adiabatic to about 90K near 42 km (2.5 mb). This last temperature represents the average temperature of the topmost 5 mb layer of the Martian atmosphere. Also shown in Figure 2 is an approximate representation of the profile computed by Goody (1957) for the same carbon dioxide content but with a surface temperature of 270K. Goody's computed tropopause is at about the same height as ours. Above the tropopause the shapes of both curves are in reasonable agreement up to about 20 mb, but above this level our temperatures decrease more rapidly than Goody's. This discrepancy may be due to a number of factors, including the low height resolution in the upper part of our model atmosphere, the boundary condition we imposed at the top of the atmosphere and the extrapolated R values used at these heights. In any case the major features of both profiles are the same:

(1) An adiabatic troposphere extending to about 9 km; and

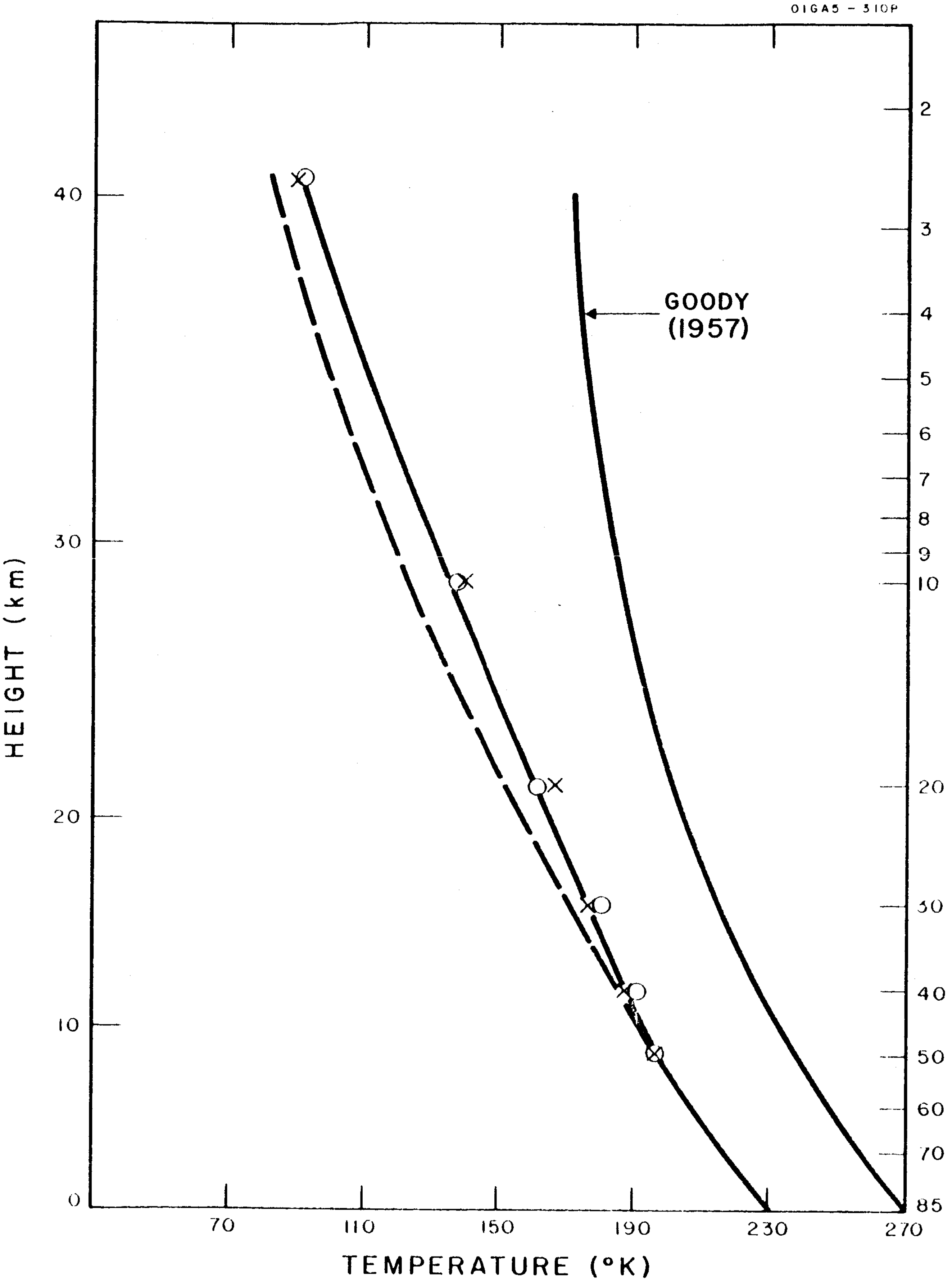

Figure 2. Computed vertical distribution of temperature in the Martian atmosphere. Crosses represent temperatures obtained from an initial isothermal state; circles represent temperatures obtained from an initial adiabatic state. Goody's (1957) profile and an adiabatic profile (dashed line) are also indicated.

(2) A stratosphere that is stable but with temperature still decreasing with height.

It is interesting to note that at the low temperatures which we have computed for the 30 to 40 km region, carbon dioxide would condense. Such a condensation might have some connection with the blue haze observed in the Martian atmosphere. However, before we can adopt our estimated temperature distribution with any degree of certainty we must first ascertain how much water vapor and ozone is actually present in the Martian atmosphere. Ozone especially, if present, would affect the stratospheric temperature distribution through its direct absorption of solar radiation. Thus, the computed temperature distribution should be considered as a tentative estimate of the true temperature distribution. Additional computations should be made as soon as better estimates of the average water vapor and ozone contents become available.

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